Oceanic control of sea level rise patterns along the East Coast of the United States

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[1] Along the eastern seaboard of the U.S. from Florida to Maine, sea level rise (SLR) shows notable patterns and significant deviation from the global mean, which have been attributed to land subsidence. Consistent with several recent studies, we analyze various observation and modeling data, and find that ocean dynamics is also an important factor in explaining these coastal SLR patterns. Despite a southward shift since the 1990s, an overall northward shift of the Gulf Stream during the twentieth century contributed to the faster SLR in the Mid-Atlantic region (North Carolina to New Jersey). In response to the 21st century climatic forcing, the rise (fall) of the dynamic sea level north (south) of Cape Hatteras is mainly induced by the significant decline of ocean density contrast across the Gulf Stream. This baroclinic process is the likely cause of the recent switch of the coastal SLR to a pattern with faster (slower) rates north (south) of Cape Hatteras.


1. Introduction

[2] Sea level rise (SLR) along the East Coast of the U.S. was faster and larger in the Mid-Atlantic region during much of the twentieth century. Long-term tide gauge (TG) records indicate that the relative SLR from North Carolina to New Jersey was about 3–5 mm yr⁻¹ (Figure 1b) [Douglas et al., 2001; Titus and Anderson, 2009], significantly faster than the global mean of about 1.7–1.9 mm yr⁻¹ [Church and White, 2011]. The SLR rates decrease toward the north and south. This middle-high coastal SLR pattern has been attributed to land subsidence mainly induced by glacial isostatic adjustment (GIA) [Boon et al., 2010; Davis and Mitrovica, 1996]. This attribution is supported by the GIA modeling [Peltier, 2004] and paleo sea level reconstructions [Engelhart et al., 2011].

[3] During the past two decades, the middle-high pattern has been replaced by a north-high (4–6 mm yr⁻¹) south-low (0–2 mm yr⁻¹) pattern, resulting from a significant SLR acceleration north of Cape Hatteras (CH) [Boon, 2012; Ezer and Corlett, 2012; Sallenger et al., 2012], most notably along the coast of New England, and a deceleration in SLR along the U.S. Southeast Coast (Figure 1b). The coastal SLR displays a step-function-like pattern with a sudden jump at CH. The local SLRs in the same SLR regime (north or south of CH) are highly correlated, suggesting a common and nonlocal mechanism. It should be noted that due to the short period, the linear sea level trend of 1993–2012 is associated with larger error bars and may not be significantly different from that of 1950–2012 at the 95% confidence level. With longer TG data and the Empirical Mode Decomposition analysis, Ezer et al. [2013] does identify a statistically significant SLR acceleration in the Mid-Atlantic region, which is attributable to the weakening of the Gulf Stream (GS). The role of the multidecadal SLR variability has also been investigated recently [Kopp, 2013]. As the SLR induced by GIA should be slow, linear, and spatially smooth, here we use ocean dynamics to understand the sharp SLR patterns along the U.S. East Coast and their rapid transition.

2. Data, Model, and Analysis Methods

[4] Various observation and modeling data of the mean sea level and sea level changes are used in the present study. For the coastal SLR, the TG data are from the Permanent Service for Mean Sea Level [Holgate et al., 2013]. Only those TG stations with long-term and high-quality records are chosen to calculate the linear trends of the relative sea level for the periods of 1950–2012 and 1993–2012 (satellite era). The eastern U.S. coastline is partitioned into 1° latitudinal bands. The average trend of all TG data in a band is calculated. To remove the land vertical movement signals, we correct the TG data with the ICE-5G model simulation [Peltier, 2004] and the Global Positioning System (GPS) observations based on the International Terrestrial Reference Frame 2005 ULR3 framework [Woppelmann et al., 2009]. Only six GPS stations are available near the TG stations along the U.S. East Coast. Another GPS data from Fernandina are not used here.

[5] In terms of the mean dynamic sea level (DSL; i.e., sea surface height relative to the geoid) observed by satellites, we use the Archiving, Validation, and Interpretation of Satellite Oceanographic (AVISO) data processed by the Jet Propulsion Laboratory (JPL) of NASA for 1993–2010 (http://podaac.jpl.nasa.gov/dataset/AVISO_L4_DYN_TOPO_1DEG_1MO). The data with 1° × 1° resolution has been recommended by the Coupled Model Intercomparison Project Phase 5 (CMIP5) for model-data comparison [Taylor et al., 2012]. We also use the 20 year (1993–2012) DSL anomaly (relative to the mean DSL) data processed by CSIRO (http://www.cmar.csiro.au/sealevel/sl_data_cmar.html) to identify the DSL variability (internal variations) and change (externally forced trend) modes in the North Atlantic. Compared to the JPL AVISO data, the CSIRO data are more up-to-date with better defined land-sea
boundary. These facilitate the calculation of coastal DSL changes and the TG-altimetry data comparison.

To study the steric contribution to the horizontal gradient of the mean DSL, we use the Polar Science Center Hydrographic Climatology (PHC) ocean potential temperature and salinity climatology with $1^\circ \times 1^\circ$ resolution and down to a depth of 5500 m [Steele et al., 2001]. The steric sea level ($h$) with respect to a reference level ($Z$) can be calculated as

$$h = -\frac{1}{\rho_0} \int_{Z}^{0} \left[ \rho(T, S, P) - \rho_0 \right] dz,$$

where $\rho_0$ is a reference density; $S$ is salinity; $T$ is in situ temperature converted from potential temperature ($\theta$); $P$ is the ocean pressure estimate based on ocean depth. In addition, the anomaly data of ocean heat content in the upper 700 m for 1955–2012 are obtained from the World Ocean Atlas [Levitus et al., 2012]. To examine the role of winds in the SLR pattern, we use the Coordinated Ocean-ice Reference Experiments forcing data to calculate the wind stress climatology and trend during 1948–2007 [Large and Yeager, 2009].

The model used in the present study is the Earth System Model (ESM2M) recently developed at the Geophysical Fluid Dynamics Laboratory of NOAA [Dunne et al., 2012]. ESM2M uses a physical climate component with a 2° atmosphere and a 1° ocean. It also incorporates component models of land/ocean ecology and biogeochemistry, thereby explicitly representing global carbon cycle. For the oceanic component, ESM2M employs the $z^*$-coordinate Modular Ocean Model (MOM4) with a free ocean surface, 50 vertical levels, and a realistic representation of the freshwater flux at the ocean surface. The DSL ($\eta$) in ESM2M is a prognostic variable. After a sufficient model spin-up period, the drift in the control run is negligible. The twentieth century simulation and 21st century projection with ESM2M have been performed subsequently according to the CMIP5 protocol [Taylor et al., 2012].

3. Results

In the current climate, the DSL is low along the eastern seaboard of the U.S. compared to the interior of the subtropical gyre (i.e., the Sargasso Sea). The steep offshore sea level slope provides the pressure gradient force to drive the strong and narrow GS. In addition, the DSL also shows a pronounced alongshore gradient. The coastal DSL gradually drops by about 0.6 m from Florida to Maine (Figures 1a and 2b). This is mainly because the oceans near the Northeast Coast of the U.S. are occupied by cold, fresh, and dense subpolar water mass (Figure 2a), which significantly lowers the steric height (Figure 2c). So the anomalously low coastal sea level north of CH is maintained by the strong buoyancy loss (density gain) in the subpolar gyre.

By contrast, warm, salty, and relatively light subtropical water mass near the Southeast Coast of the U.S. leads to a higher coastal DSL. Superimposed on the general northward gradient, the coastal DSL displays a local high around CH due to the convergence of the northward GS and the southward Slope Current (Figure 1a). As a result, the alongshore DSL gradient is particularly strong north of CH. Both the offshore and alongshore DSL gradients are well captured.

![Figure 1. Mean dynamic sea level (DSL) and SLR patterns along the East Coast of the U.S. (a) Observed (red curve; mean of 1993–2010) and simulated (blue curve; mean of 1986–2005) DSL. The rectangles indicate the cross-current DSL difference ($\Delta \eta$; high values minus low values) along the green lines shown in Figure 2b. The simulations are from the twentieth century run of ESM2M. The latitudes of the eastern U.S. coastal cities are also indicated. (b) Coastal SLR rates (linear trend) from the TG data for the periods of 1950–2012 and 1993–2012. The raw TG data are corrected on the basis of both the GIA modeling and GPS observations about the vertical land movement. (c) Decadal trends (1993–2002 and 2003–2012) of the Commonwealth Scientific and Industrial Research Organisation (CSIRO) altimetry data, together with the 20 year overall trend. The error bar indicates $\pm$ one standard error. Notice that the decadal trend is sensitive to the start and end years.](image-url)
in the twentieth century simulation of the ESM2M model (Figure 1a). But the model misses the local high around CH due to its relatively coarse resolution (1° × 1°).

[10] The mean coastal DSL is closely related to the DSL difference across the GS (Δη). Δη is about 0.2–0.5 m upstream of CH but increases significantly to 0.9–1.1 m downstream of CH (Figure 1a). The abrupt increase in Δη across the GS separation is a result of the enhanced cross-current ocean density contrast and stronger baroclinicity. Downstream of CH, the GS serves as the front to separate the distinct subpolar and subtropical water masses (Figure 2a) [Bower et al., 1985]. The steric effect in the upper 1500 m accounts for about 85% of the 1 m DSL difference between Points C and D (Figures 2b and 2c). Especially, the thermosteric effect associated with the cross-current temperature contrast generates 1.70 m sea level difference (Figure S1 in the supporting information), which is partially compensated by the halosteric effect (−0.86 m; the negative sign indicates a reverse gradient). The smaller Δη upstream of CH (0.40 m between Points A and B) is due to similar waters found on both sides of the GS and a much weaker steric height gradient. Therefore, Δη downstream of CH and the alongshore DSL gradient are mainly due to the temperature variation associated with the distribution of different water masses.

[11] Given the significant pattern of the mean DSL, its modification can cause nonuniform SLR. During 1993–2012, the north-high south-low SLR pattern along the U.S. East Coast remains pronounced in the TG data even after the GIA corrections (Figure 1b). The altimetry data shows a relatively smooth coastal SLR pattern during the past two decades (Figure 1c). The decadal trends of the altimetry data indicate that the north-high south-low SLR pattern is strengthening during the 21st century (Figures 1c, 3g, and 3h). Whether it reflects a short- or long-term behavior is unclear due to the insufficient data length. This is because both the multidecadal variability and the long-term trend are important in the North Atlantic [Ezer et al., 2013; Kopp, 2013; Wunsch, 2013]. Unlike the point measurement by TG, the altimetry data represent the DSL averaged over a wide ocean area. The discrepancy between the two data sets may be also due to sampling, land influence, and geophysical correction issues in the coastal application of the altimetry data [Bouffard et al., 2008].

[12] An empirical orthogonal function (EOF) analysis on the 20 year annual mean altimetry data reveals that the DSL in the North Atlantic is dominated by a covarying dipole (Figure 3a) [Hakkinen and Rhines, 2004; Li et al., 2012; Lorbacher et al., 2010; Zhang, 2008]. The positive center coincides with the axis of the GS and stretches from the
The ocean interior all the way to the U.S. Mid-Atlantic coast. The negative center is located at the eastern subpolar gyre. ESM2M simulates a similar DSL dipole during the twentieth century (Figure 3b), although the positive region does not strongly intercept the Mid-Atlantic coast and the negative center extends to the south. The principle component (PC1) of this dipole shows pronounced decadal to multidecadal variability (Figure 4a) and is highly correlated with the index of the Atlantic meridional overturning circulation (AMOC). The correlation coefficient is $-0.8$ for 4-year lag (Figure S2), suggesting a robust AMOC-DSL relationship linked by the northward heat transport variation in the upper

Figure 3. Two modes of the DSL variability and change in the North Atlantic. (a) The leading EOF of the CSIRO altimetry data for 1993–2012 (shading) and the observed mean DSL (JPL AVISO; contours; m). (b) The leading EOF of the DSL in the historical run (1861–2005) of ESM2M (shading) and the simulated mean DSL (contours). (c) The leading EOF of the observed ocean heat content in the upper 700 m for 1955–2012 [Levitus et al., 2012] and the observed mean DSL (contours). (d) The linear trend of wind stress (N m$^{-2}$ yr$^{-1}$) and the Sverdrup stream function (Sv yr$^{-1}$) for the period of 1948–2007 [Large and Yeager, 2009]. Positive stream function values indicate clockwise circulation. (e) The leading EOF of the DSL in the projection run (2006–2100) of ESM2M under the RCP8.5 greenhouse-gas emission scenario (shading) and the simulated mean DSL during the twentieth century (contours). (f) Projected anomalies of ocean surface buoyancy flux ($10^{−8}$ m$^{2}$ s$^{-3}$) and wind stress (N m$^{-2}$) during 2091–2100 relative to 1986–2005. Positive values indicate reduction of the buoyancy loss. (g and h) Decadal trends (mm yr$^{-1}$) of the CSIRO altimetry data for periods of 1993–2002 (Figure 3g) and 2003–2012 (Figure 3h). Contours are the observed mean DSL (m). The EOF analysis is based on the annual mean data with the global long-term trend removed.
Figure 4. Principle components of the two (twentieth and 21st century) DSL modes and the related changes. (a) PC1 of the observed, simulated, and projected DSL, as well as the observed ocean heat content in the upper 700 m and the simulated and projected AMOC index by ESM2M (defined as the maximum overturning stream function at 45°N in the North Atlantic). Thin lines indicate the linear trend. (b) ESM2M simulated and projected area mean buoyancy flux and the northward heat transport at 45°N in the Atlantic. (c) ESM2M simulated twentieth century mean DSL at Line C-D and the anomalies induced by a unit change of the PC1 of the twentieth and 21st century modes as shown in Figure 4a.

Atlantic (Figure 4b) [Ezer et al., 2013; Sallenger et al., 2012]. Ocean temperature and density variations associated with the dipole are confined within the GS, especially in the subsurface layer (Figure S3). Indeed, the observed ocean heat content and thermosteric SLR in the upper 700 m display a similar dipole pattern since the 1950s (Figures 3c and 4a), reflecting redistribution of ocean heat between the GS and subpolar gyre. So the positive (negative) phase of the DSL dipole is closely related to the weakening (strengthening) of the AMOC on decadal to multidecadal timescales, heat convergence (divergence) in the GS, and a northward (southward) shift of its route downstream of CH [Joyce and Zhang, 2010].

During the twentieth century, the DSL dipole in the ESM2M simulation displays a long-term upward trend superimposed on its decadal to multidecadal variability (Figure 4a). This is consistent with a gradual weakening of the AMOC. In addition, the westerlies showed a northward shift during the twentieth century (Figure 3d), which caused relatively large Sverdrup stream function anomalies near the Mid-Atlantic, a northward shift of the GS [Miller and Douglas, 2004], and hotspots of surface ocean warming [Wu et al., 2012]. Compared to other western boundary currents, the GS region is unique in the net heat accumulation in the upper 700 m (Figure S4). The rapid warming and northward shift of the GS from the 1950s to early 1990s was followed by a cooling and southward shift since 1993 (note that Ezer et al. [2013] found a southward shift near the Mid-Atlantic and a weakening of the GS during the last decade).

According to Gill [1982], the buoyancy flux \( B \) (unit: \( \text{m}^2 \text{s}^{-3} \)) can be calculated as

\[
B = \left[ g \rho_0 \left( \alpha Q_{HF} + \beta P - \rho_0 S (E - P - R) \right) \right],
\]

where \( g \) is the gravitational acceleration; \( \rho_0 \) is the reference density; \( \alpha \) is the specific heat for seawater; \( \beta \) is the ocean surface salinity; \( \alpha \) and \( \beta \) are the thermal expansion and saline contraction coefficients, respectively; \( Q_{HF} \) is the net air-sea heat flux; and \( E, P, \) and \( R \) are evaporation, precipitation, and runoff, respectively. The buoyancy loss in the subpolar gyre remains strong in the twentieth century ESM2M simulation (Figure 4b). The continuing formation and southward spread of the subpolar water mass in the northwestern North Atlantic would keep the ocean front along the GS sharp. It should be noted that all of these processes are inherently linked in the coupled climate system. Given that the Mid-Atlantic is just north of the GS separation, the coastal sea level in this region is sensitive to shifts of the GS position and changes in its strength [Ezer et al., 2013]. In other words, if the red curve in Figure 1a shifts northward, it results in faster and larger SLR in the Mid-Atlantic region.

In the 21st century projection under the RCP8.5 scenario, the DSL in ESM2M switches to another mode of variability and change (Figure 3e). Here we consider a strong forcing scenario to facilitate identification of the signal. This 21st century mode is characterized with a rise (fall) of the DSL north (south) of the GS, and a north-high south-low pattern of SLR along the U.S. East Coast [Yin, 2012; Yin et al., 2009]. During the 21st century, the buoyancy loss in the subpolar gyre and the AMOC weakens significantly (Figures 3f and 4b), leading to a large ocean density decrease north of the GS, and a great decline of the cross-current density contrast downstream of CH (Figure S3). \( \Delta \rho \) declines by 0.35 m at Line C-D along 65°W, 83% (0.29 m) of which is due to the steric effect and baroclinic process.
associated with the subpolar water mass modification in the northwestern North Atlantic (Figures 4c, S5, and S6). The DSL change can be decomposed into a baroclinic and a barotropic component:

\[ g\Delta \eta = \frac{g}{H_0} \int \left[ V_0' \delta \, dz - f \mathbf{k} \times \mathbf{V}' \right] \, dz, \]

where \( \eta' \), \( \rho' \), and \( \mathbf{V}' \) are the anomalies of the DSL, ocean density, and deep-mean barotropic velocity, respectively; \( f \) and \( H \) are the Coriolis parameter and ocean depth, respectively; and \( \mathbf{k} \) is the vertical unit vector [Low and Gregory, 2006]. The first term on the right-hand side of equation (3) is the baroclinic part due to the change of the density contrast across the GS (i.e., the steric effect). The second term is the barotropic contribution related to the change of the GS transport.

By contrast, \( \Delta \eta \) declines by only 0.13 m at Line A-B along 30.5°N, with small contribution from the steric effect. The barotropic process must be responsible for the DSL change near the Southeast Coast of the U.S. Over the 21st century, ocean volume or mass transport by the GS in the upper 1000 m shows similar reductions upstream and downstream of CH, or by about 12 and 13 Sv at Lines A-B and C-D, respectively. A simple calculation based on the geostrophic relationship indicates that the barotropic process associated with these reductions can lead to 0.8–0.10 m decline of \( \Delta \eta \). So through the barotropic process only, the cross-current DSL gradient is relatively insensitive to the change of the GS transport, which is the case upstream of CH and near the U.S. Southeast Coast. Downstream of CH, by contrast, the baroclinic process is much more efficient at modifying \( \Delta \eta \) and is mainly responsible for the faster and larger SLR along the U.S. Northeast Coast.

4. Summary

Our results suggest that ocean dynamics is an important factor in causing the spatiotemporal characteristics of SLR in the North Atlantic and along the East Coast of the U.S. During the twentieth century, the variability and change of the DSL in the North Atlantic was dominated by a dipole between the GS and subpolar gyre. The middle-high DSL pattern along the U.S. East Coast results, at least in part, from shifting in the GS location and its strength induced by changes in the winds, AMOC, and the North Atlantic Oscillation [Ezer et al., 2013]. When external climatic forcing passes some critical level, the coastal SLR switches to the north-high south-low pattern during the 21st century. It is mainly induced by the thermohaline process related to diabatic heating and ocean freshening. The wind change plays a secondary role in this mode (Figure 3f) [Yin et al., 2010]. The high (low) sensitivity of the coastal DSL north (south) of CH to the AMOC stems from the dominance of baroclinic (barotropic) processes, with the former much more effective at altering the DSL gradient. Our results indicate that the DSL in the North Atlantic and along the U.S. East Coast could be an important fingerprint of the AMOC variability and change. Given that the deep (dense) water formation in the Labrador and Irminger Sea is sensitive to climatic warming and the melting of the Greenland ice sheet [Hu et al., 2009; Yin et al., 2009], the Northeast Coast of the U.S. is particularly vulnerable to future SLR and associated coastal hazards and disasters such as the elevated storm surge induced by Superstorm Sandy.

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